CLIMATE IN THE DRY CENTRAL ANDES OVER GEOLOGIC, MILLENNIAL, AND INTERANNUAL TIMESCALES[1]

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Abstract

Over the last eight years, we have developed several paleoenvironmental records from a broad geographic region spanning the Altiplano in Bolivia (18°S–22°S) and continuing south along the western Andean flank to ca. 26°S. These records include: cosmogenic nuclide concentrations in surface deposits, dated nitrate paleosoils, lake levels, groundwater levels from wetland deposits, and plant macrofossils from urine-encrusted rodent middens. Arid environments are often uniquely sensitive to climate perturbations, and there is evidence of significant changes in precipitation on the western flank of the central Andes and the adjacent Altiplano. In contrast, the Atacama Desert of northern Chile is hyperarid over many millions of years. This uniquely prolonged arid climate requires the isolation of the Atacama from the Amazon Basin, a situation that has existed for more than 10 million years and that resulted from the uplift of the Andes and/or formation of the Altiplano plateau. New evidence from multiple terrestrial cosmogenic nuclides, however, suggests that overall aridity is occasionally punctuated by rare rainfall events that likely originate from the Pacific. East of the hyperarid zone, climate history from multiple proxies reveals alternating wet and dry intervals where changes in precipitation originating from the Atlantic may exceed 50%. An analysis of Pleistocene climate records across the region allows reconstruction of the spatial and temporal components of climate change. These Pleistocene wet events span the modern transition between two modes of interannual precipitation variability, and regional climate history for the Central Andean Pluvial Event (CAPE; ca. 18–8 ka) points toward similar drivers of modern interannual and past millennial-scale climate variability. The north-northeast mode of climate variability is linked to El Niño–Southern Oscillation (ENSO) variability, and the southeast mode is linked to aridity in the Chaco region of Argentina.

Key words: Altiplano, Amazon Basin, Andes, CAPE, ENSO, middens.

The dry central Andes is the tripartite region encompassing the Altiplano, the Atacama, and the western Andean flank between ca. 18°S and 27°S (Fig. 1) and is a critical region for understanding the drivers of tropical climate change at multiple timescales. Arid environments are often uniquely sensitive to climate change, and today modern interannual climate variability in the region is pronounced and influenced by both tropical climate phenomena, such as El Niño–Southern Oscillation (ENSO), and moisture levels in the extratropical lowlands east of the Andes (Fig. 2) (Vuille & Keimig, 2004). Here, we compare the timing and likely drivers of precipitation changes across the dry central Andes over geologic, millennial, and interannual timescales. Understanding how regional climate is sensitive to processes like mountain building, the reorganization of global atmospheric circulation that occurs over glacial-interglacial and millennial timescales, and decadal to interannual changes due to processes such as the ENSO phenomenon is a step toward assessing where and how this region is sensitive to global climate change.

The Atacama Desert, located along the western Andean slope between ca. 18°S and 26°S (Fig. 1), is

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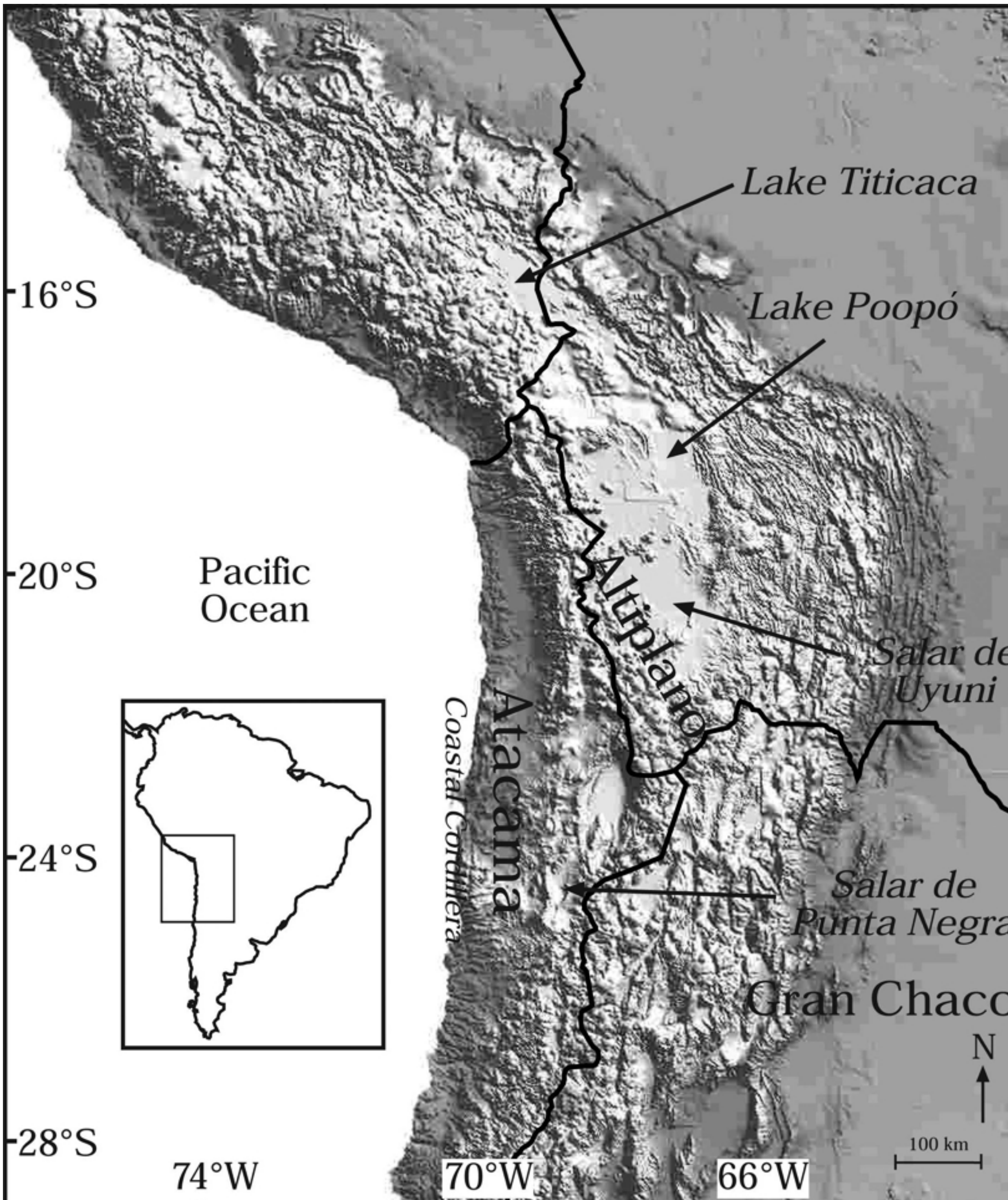

Figure 1. Location of relevant sites and geographic features in the dry central Andes.

the driest and perhaps oldest desert on earth (Hartley et al., 2005). Hyperaridity requires orographic exclusion of Atlantic moisture by the Andes and exclusion of Pacific moisture by the Coastal Cordillera and subsiding air resulting from the cold, northward-flowing Humboldt Current. The stability and timing of moisture exclusion from these two sources are critical to determining if the Andean uplift created the Atacama or if this aridity results from changes along the Pacific coast (Lamb & Davis, 2003). Despite this prolonged aridity, major changes have occurred in the boundary conditions that contribute to hyperaridity since the Andes acquired enough elevation to constitute a significant orographic barrier. These changes include uplift of the Coastal Cordillera (e.g., Clift & Hartley, 2007) and changes in the intensity of the Humboldt Current (e.g., Molnar & Cane, 2007) related to expansion of Antarctic ice

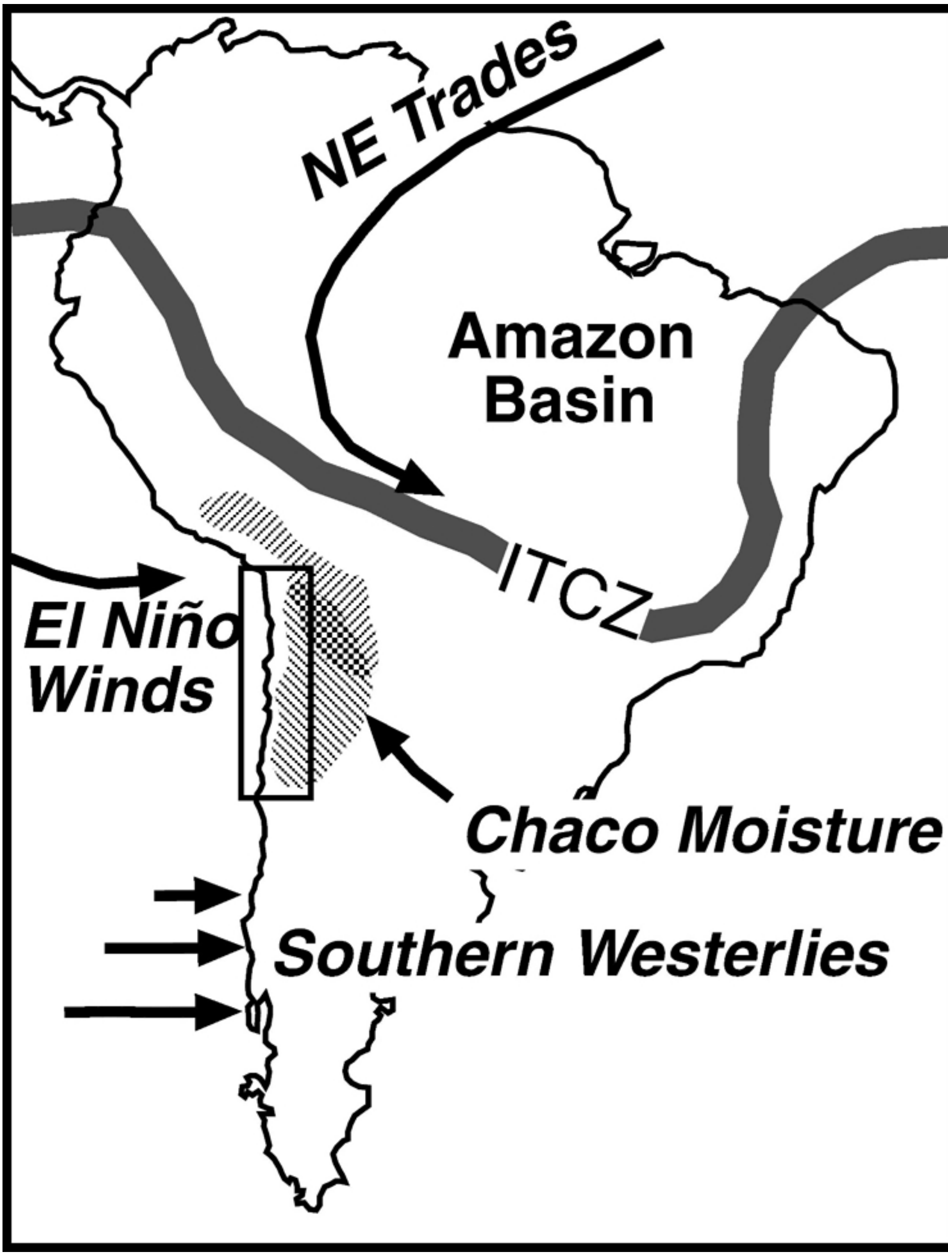

Figure 2. Modern climate systems controlling central Andean rainfall. Overlapping shaded zones show the two modes of modern precipitation, as major rotated empirical orthogonal functions, identified by Vuille and Keimig (2004). The north-northwest mode is modulated by El Niño–Southern Oscillation (ENSO), with strong westerly winds producing drought on the Altiplano during El Niño years. The southeast mode is correlated with lowland humidity in the Chaco region of Argentina. The Intertropical Convergence Zone (ITCZ) is shown in its southernmost (summer) position.

sheets (e.g., Hartley & Chong, 2002) and/or closing of the Isthmus of Panama (Ibaraki, 1997).

The evidence for prolonged hyperaridity in the core of the Atacama Desert is matched by paleoecological (e.g., Grosjean et al., 1997; Betancourt et al., 2000; Latorre et al., 2002, 2006) and paleohydrological (e.g., Betancourt et al., 2000; Bobst et al., 2001; Rech et al., 2002; Quade et al., 2008) evidence for dramatic millennial scale changes in climate along the fringes of the Atacama Desert. Thus, the boundaries of the Atacama Desert fluctuate in response to these climatic events, and the distribution and stability of these boundaries through time can give insights into the causes of these shifts. Recent evidence (e.g., Quade et al., 2008) suggests that ancient millennial scale variability had two geographically distinct modes, similar in distribution to two distinct modes of modern interannual rainfall variability.

Atacama Hyperaridity

The hyperaridity of the Atacama Desert is due to a combination of: (1) the extreme rain shadow created by the high Andes and Altiplano, which excludes moisture from the Amazon Basin; (2) a strong temperature inversion along the Pacific coast, which effectively blocks Pacific moisture at ca. 1000 m elevation along the western flank of the Coastal Cordillera; and (3) the northern limit of the southern Westerlies (Houston & Hartley, 2003). Over millions of years, the rain shadow created by the high Andes and/or Altiplano plateau was primarily responsible for the prolonged aridity of the Atacama Desert. An Andean elevation of at least 2000 m is considered high enough to exclude much of the moisture originating in the Amazon Basin from the Atacama (e.g., Masek et al., 1994; Rech et al., 2006). The details and timing of central Andean uplift and formation of the Altiplano plateau and the interaction between climate and tectonics in the central Andes remain unresolved (e.g., Barnes et al., 2006; Garzione et al., 2006; Ghosh et al., 2006).

One of the primary lines of evidence for both a landscape and climate that has remained stable and hyperarid over the entire Pliocene and Pleistocene is extremely high cosmogenic nuclide concentrations from ancient geomorphic surfaces. Cosmogenic nuclides are produced by secondary cosmic rays in the uppermost few meters of the earth's surface and can record the age of material suddenly exposed or constrain erosion rates (Lal, 1991). Cosmogenic nuclide concentrations from stable geomorphic surfaces in the Atacama result in some of the oldest exposure ages found anywhere on earth, ranging between 9 and 37 million years ago (Ma) (Dunai et al., 2005; Nishiizumi et al., 2005; Kober et al., 2007). Indeed, the Atacama is one of the few locations where exposure ages must be verified by stable ^{21}Ne measurements, as long exposure times mean that significant quantities of the radionuclides ^{10}Be and ^{26}Al produced during early exposure have decayed. Constraints on the rates of sediment production and transport in the Atacama also come from cosmogenic nuclide concentrations in multiple components of the landscape (Placzek et al., 2007) and deposition rates inferred from dated ash-fall tuffs (Placzek et al., 2009). Together, these indicate that overall erosion rates are some of the slowest in the world—a direct result of a prolonged arid climate.

Additional evidence for the onset of aridity prior to 10 Ma includes: accumulation of nitrate soils in ancient deposits (Rech et al., 2006), an end of supergene mineralization (e.g., Alpers & Brimhall, 1988; Sillitoe & McKee, 1996; Arancibia et al., 2006), and changes in stream morphology on the Andean flank (Hoke et al., 2006). Ancient nitrate soils, with a firm minimum age of 9.4 Ma from an overlying volcanic ignimbrite, attest to this ancient aridity as nitrates require hyperarid conditions and today only accumulate in the driest portions of the Atacama Desert. These nitrate soils, however, probably represent several million years of accumulation, and Rech et al. (2006) place the minimum age for the onset of hyperaridity at ca. 13 Ma.

At odds with all this evidence for prolonged hyperaridity is an inferred association between the degree of aridity and the deposition of fluviolacustrine, alluvial fan or evaporite deposits, which leads to the conclusion that Pliocene sediments suggest a transition from arid to hyperarid conditions as recently as 3 Ma (Hartley & Chong, 2002; Allmendinger et al., 2005). Today, all of these depositional environments occur both in the wetter Andean highlands and across the "absolute desert," a broad expanse of the Atacama Desert completely devoid of precipitation and vascular plants, thus confounding interpretation of modern or ancient aridity from such sediments. Cosmogenic nuclide concentration from the active components of the landscape (surface gravel, active alluvial fan deposits, and active channels) appears to be eroding at a rate that is at least an order of magnitude faster than relict geomorphic surfaces (Placzek et al., 2007). Furthermore, new ^{21}Ne, ^{10}Be, and ^{26}Al measurements from relict boulders indicate that many of these boulders have ages less than 3 Ma (Placzek et al., 2008), long after the onset of aridity. This movement and erosion of all size classes of sediment after 3 Ma suggest that periodic rainfall and flood events continue to impact the Atacama. Furthermore, it suggests that the Atacama Desert, traditionally viewed as isolated from rainfall over geologic intervals, has a modern landscape that is shaped by rare, but recent, rain events and is therefore not fully isolated from future global climate change.

Millennial-scale Climate Change

Climate proxies from lakes, wetland deposits, and urine-encrusted rodent middens reveal dramatic precipitation changes throughout the Pleistocene over a broad geographic region. Here, we focus on the paleolake record from the Altiplano and what it reveals about climate variability over the Pleistocene. We also compare this lake record to other types of climate proxies across this region during the post late glacial–age Central Andean Pluvial Event (CAPE), concluding with an example of how a multiproxy approach allows tracking of the source of moisture during wet intervals.

LAKE RECORDS

Four large lake basins (Fig. 1: Titicaca, Poopó, Coipasa, and Uyuni) dominate the Altiplano, and the size of the lakes has undergone periodic changes as a result of changes in precipitation. In the north, Lake Titicaca (3806 m elevation, 8560 km^2) is a freshwater lake that is more than 284 m deep (Argollo & Mourguiart, 2000) and loses less than 10% of its annual water budget to overflow into the Río Desaguadero (Roche et al., 1992). The Río Desaguadero empties into the oligosaline Lake Poopó (3685 m, 2500 km^2), which is separated by a topographic divide, the Laka sill (3700 m), from the Salar de Coipasa (3656 m, 2530 km^2) and Salar de Uyuni (3653 m, 12,100 km^2). In wet years these salt flats are connected and filled with shallow water (< 4 m) (Argollo & Mourguiart, 2000).

Within these basins, multiple sites were studied and sampled as part of a comprehensive effort to obtain and replicate records of lake-level change from multiple localities in all three major basins. Particular effort was directed toward sedimentary deposits associated with various visible paleoshorelines. This approach to reconstructing lake-level history allows for direct determination of lake level, replication of stratigraphy, and dating by two geochronologic methods (^{14}C and U-Th, Placzek et al., 2006b). More than 170 dates are available from paleolake deposits within the basins, and the use of both the U-Th and radiocarbon methods allowed us to extend our record beyond the limit of radiocarbon dating (ca. 45 ka). The focus of this dating effort is sedimentary deposits indicative of a near-shore environment and the massive encrustations of calcium carbonate (tufas) found in the paleolake basins. Tufas and aquatic gastropod shells generally form in nearshore environments and incorporate ^{14}C and uranium from water in which they precipitate. For samples younger than 45 ka, the quantity of remaining radioactive ^{14}C can be used to calculate a sample's age. For older samples (greater than 25 ka), however, the very small quantity of remaining ^{14}C renders samples susceptible to errors introduced by contamination with very small amounts of modern carbon. Thus, reliable ages greater than 25 ka come from the U-Th dating method, which is based on the premise that uranium is incorporated into carbonates precipitated from water, but thorium, a daughter of uranium decay, is largely not incorporated into tufas. Sediments that are clearly associated with lake shorelines or sedimentary units showing both deep and shallow lake events place constraints on absolute paleolake elevation. The potential incompleteness of any single exposure is redressed by replication of stratigraphy at multiple locations (Placzek et al., 2006a).

On the Altiplano, extensive natural exposures reveal evidence of two deep-lake and several minor-lake cycles over the past 120 ka (Fig. 3) in an area where today there are mostly barren salt flats or shallow saline lakes. The Ouki lake cycle was ca. 80 m deep, and 19 U-Th dates place this deep-lake cycle between 120 and 98 ka (Placzek et al., 2006a). Old shoreline and sedimentary deposits from the Ouki lake cycle are extensively exposed in the Poopó Basin, but no deep lakes are apparent in the subsequent record between 98 and 18.1 ka. Evidence of shallow lakes is present in the Uyuni Basin between 95 and 80 ka (Salinas lake cycle), at ca. 46 ka (Inca Huasi lake cycle), and between 24 and 20.5 ka (Sajsi lake cycle) (Fig. 3). The Tauca lake cycle occurred between 18.1 and 14.1 ka, resulting in the deepest (ca. 140 m) and largest lake in the basin over the past 120 ka. Multiple ^{14}C and U-Th dates constrain the highest elevation of the Tauca lake cycle along a topographically conspicuous shoreline between 16.4 and 14.1 ka. The Coipasa lake cycle produced a $\leq$ 55 m deep lake with ages between ca. 13 and 11 ka (Placzek et al., 2006a). Together, the Tauca and Coipasa lake cycles evidence the occurrence of CAPE on the Bolivian Altiplano (Fig. 4).

RODENT MIDDENS

Urine-encrusted rodent middens (henceforth, rodent middens) are complex nests of local vegetation and feces encased in crystallized rodent urine. In arid climates, rodent middens are preserved underneath rock slabs and within caves. Plant remains encased in middens reflect former vegetation cover within the rodent's foraging range, which is usually less than 200 m (cf. Salinas & Latorre, 2007). In the dry central Andes, middens are produced by at least four different rodent families: Abrocomidae (*Abrocoma cinerea* Thomas, chinchilla rats), Chinchillidae (*Lagidium viscacia* Molina and *Lagidium peruanum* Meyen, southern mountain viscacha), Muridae (*Phyllotis* spp., leaf-eared mice), and Octodontidae (*Octodontomys gliroides* Gervais & d'Orbigny [1884], mountain degu [Latorre et al., 2005]). These rodents collect plants for consumption and nest building, and studies of modern *Phyllotis*, *Lagidium*, and *Abrocoma*

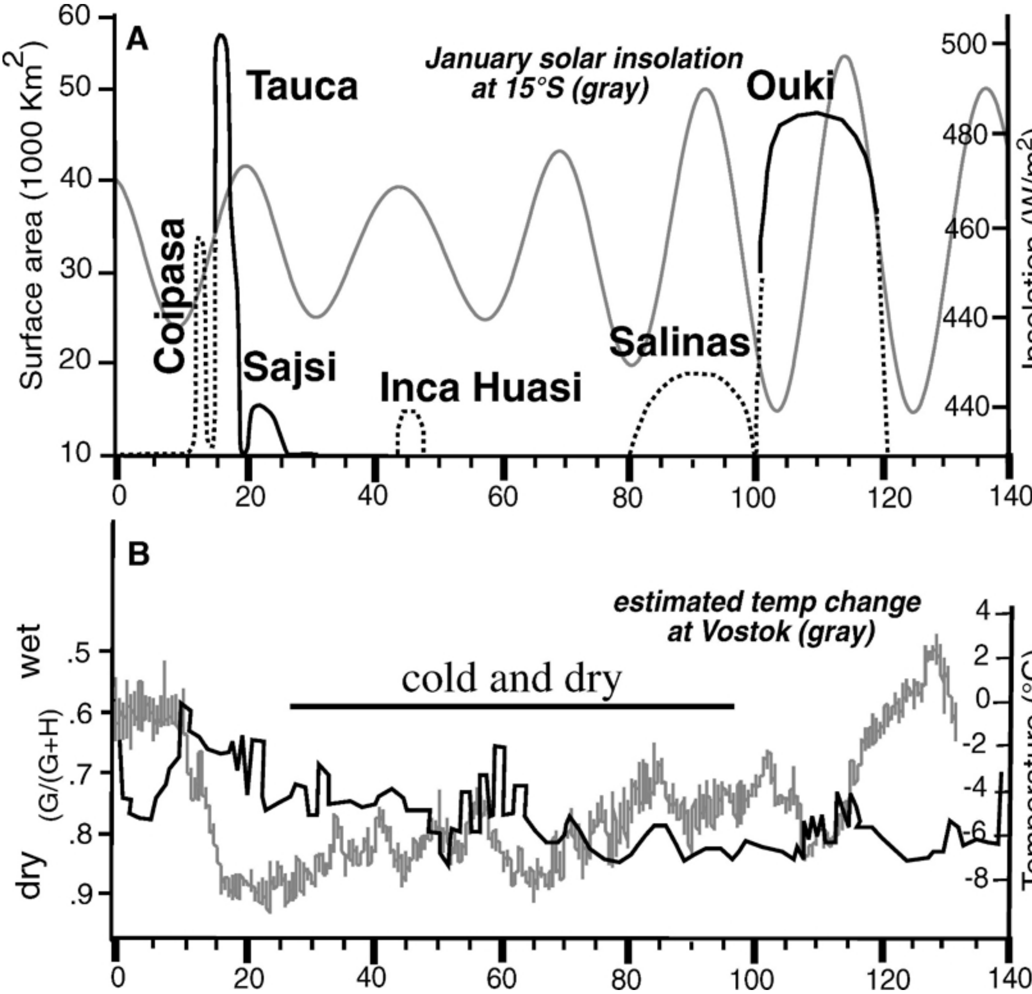

Figure 3. —A. Reconstructed lake history from shoreline deposits. January insolation (in watts/m^2) at 15°S (Laskar, 1990) is given in gray. —B. Iron oxide composition (goethite/[goethite + hematite]) of sediments derived from the Amazon (Harris & Mix, 1999) and estimated temperature change at Vostok (gray) (Petit et al., 1999). X axis values denote time in ka.

indicate that they are dietary generalists (cf. Cortés et al., 2002), and as such they are not likely to introduce large selective biases into the midden record.

Due to the abundance of plant macrofossils, rodent middens are rich snapshots of local paleoecology at the finite (and datable) time they were deposited. Rodent middens deposited within the last 45 ka are dated using standard ^{14}C techniques. Analysis of ancient vegetation assemblages is most effective when coupled with surveys of modern vegetation in and around a collection site. The most basic analysis of rodent middens typically involves assessment of the percent of extra-local plant species contained in a midden and some interpretation of the relative climate (wetter, dryer, warmer, colder) represented by that assemblage. At the outer edges of the Atacama Desert, middens containing abundant vegetation are found on landscapes that are currently too dry to support plants (Betancourt et al., 2000; Latorre et al., 2002). A simple proxy for precipitation amount from the central Andean midden record is the relative abundance of grass, as modern grasslands are currently found where there is higher precipitation present in fossil middens near the boundary of the Atacama Desert (Latorre et al., 2003, 2005, 2006). The percentage of grass abundance from rodent middens on the fringes of the absolute desert in the Salar de Punta Negra region is generally high during the CAPE (Latorre et al., 2002). Here, rodent middens from the second phase of CAPE have a higher percentage of grass abundance than the first phase of CAPE (Fig. 4).

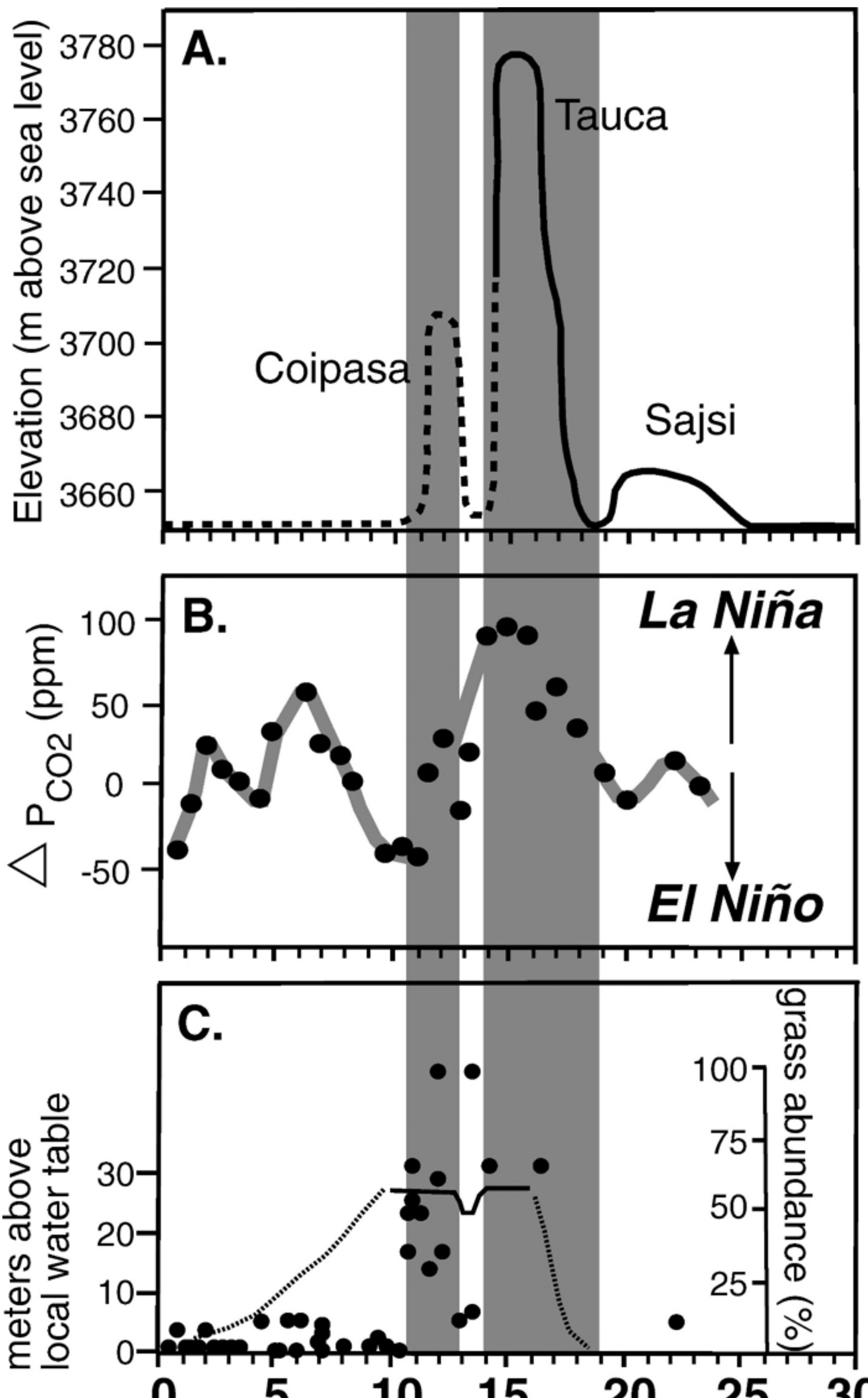

Figure 4. Comparison of paleohydrologic and climate proxies during the Central Andean Pluvial Event (CAPE). —A. Reconstructed lake-level curve. —B. Change in pCO_2 in the Western Equatorial Pacific inferred from boron isotope analyses of planktonic foraminifera, in which increased pCO_2 is associated with stronger upwelling and La Niña–like conditions (Palmer & Pearson, 2003). —C. Reconstructed water table height (Quade et al., 2008) and percentage of grass abundance from rodent middens in the Salar de Punta Negra area (Latorre et al., 2002). X axis values denote time in ka.

Paleowetlands

Wetlands form where the water table intersects the land surface and can be found either within steep-walled washes or in less confined settings where small local closures allow pooling of shallow freshwater and the formation of wetland deposits (Rech et al., 2002, 2003; Grosjean et al., 2005; Quade et al., 2008). Paleowetland deposits generally consist of fine sand, silt, and biogenic deposits such as organic-rich mats, diatomites, and tufa. The abundance of organic material in these deposits makes them relatively easy to date using radiocarbon, and multiple stratigraphic levels within a deposit can often be dated. Furthermore, the abundance of these deposits in the Atacama allows replication of results both within and between sites. Questions regarding hydrologic response time can be resolved by comparison of wetlands from several different settings; in the Atacama we find that increased precipitation in the high Andes is very rapidly translated into water table rise at multiple locations across the Atacama (Rech et al., 2002, 2003; Quade et al., 2008). High water tables in the Salar de Punta Negra region indicate that CAPE began in this region at ca. 17 ka, but may have terminated as late as 8 ka (Fig. 4).

Spatial and Temporal Extent of the CAPE

Evidence from the CAPE is relatively recent and well preserved, allowing evaluation of the spatial and temporal distribution of climate change over the entire dry central Andes. The CAPE is divided into two phases, and the depths of the Tauca and Coipasa lake cycles suggest that the first phase of CAPE on the Altiplano was the wettest and began at ca. 18 ka. This contrasts with climate records from wetlands in the Punta Negra region (ca. 4°S of the Uyuni Basin), where high water tables indicate that the second phase of CAPE began ca. 1000 years after the transgression of Lake Tauca. In both areas, the first phase of CAPE terminates abruptly at ca. 14.1 ka and is soon followed by a second wet interval (Fig. 4). The second phase of CAPE created the shallower Lake Coipasa on the Altiplano, but the midden record from the Salar de Punta Negra region has a higher percentage of grass abundance during the second phase of CAPE, an indication that this second phase was wetter toward the south (Latorre et al., 2002). While the termination of both the Coipasa lake cycle and CAPE in the Punta Negra region is poorly constrained in time, the second phase also seems to be longer lived to the south (Quade et al., 2008).

Paleolake shoreline evidence from the Altiplano also supports the assertion that the Coipasa lake cycle was sustained mainly from precipitation in the southern Coipasa and Uyuni basins. Climate affects lake levels in closed basins by altering the hydrologic balance between runoff, precipitation, and evaporation while basin topography influences lake levels by altering the surface area:volume ratio. In large lake systems elsewhere (e.g., Bonneville, Lahontan, Lisan), well-developed shorelines correspond to periods when a lake level was stabilized as a result of spilling over into an arid receiving basin at a lower level (Curry & Oviatt, 1985; Benson & Paillet, 1989; Benson et al., 1990; Bartov et al., 2002). Thus, a lake system is buffered to climate fluctuations at the level of a spillway because the receiving basin must fill before the lake in the spillover basin can again rise. The degree of buffering depends on the relative size of the two basins. In the case of the Poopó-Coipasa-Uyuni system, the Poopó Basin is considerably smaller (< 1/3 the size) than the combined Coipasa-Uyuni basins (Fig. 5). Thus, if a lake filled these basins with water from the north (the Titicaca and Poopó basins), then such a lake would have a relatively long period of stability at the level of the Laka sill (the spillway between Poopó and Coipasa). This should result in a prominent shoreline in the Poopó Basin at ca. 3700 m, the elevation of the Laka sill. In contrast, if a lake filled the larger and more southern Coipasa and Uyuni basins first, then the percentage of change in surface area at the level of the Laka sill is much smaller, so pronounced shorelines would not develop (Fig. 5). The maximum elevation of the Coipasa lake cycle remains poorly constrained because a prominent shoreline is not visible. Chronological evidence, however, suggests that at its maximum extent the Coipasa lake cycle approximated the elevation of the Laka sill.

Modern Climate Variability

Today, the sources, timing, and variability of precipitation are different for the northern Altiplano, the southern Altiplano, western Andean flank, and the Atacama. On the northern Altiplano, more than 80% of total annual precipitation falls in the austral summer (December to March) (Vuille, 1999), and this moisture traverses the Amazon Basin in the summer months when the Intertropical Convergence Zone (ITCZ) is displaced southward and convection is most intense in the Amazon Basin (Lenters & Cook, 1997) (Fig. 2). This moisture source to the north and east of the Altiplano produces a pronounced north-south gradient and is referred to as the South American Summer Monsoon (SASM) (e.g., Zhou & Lau, 1998). The SASM on the northern Altiplano is modulated by ENSO variability, and the strength of the trade winds

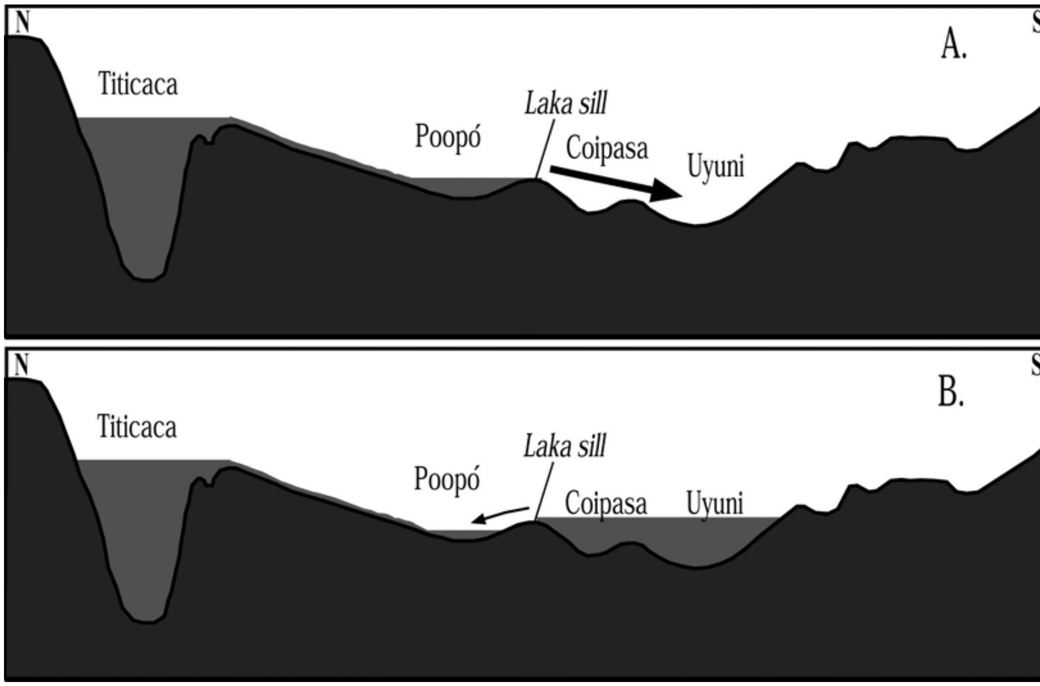

Figure 5. Schematic cross section of the Titicaca-Poopó-Coipasa-Uyuni hydrographic basin. Vertical exaggeration is ca. 830×. —A. Filling of the basins from the north. —B. Filling of the basins largely from the south.

is strong in La Niña years, resulting in increased precipitation. Conversely, during El Niño years, aridity dominates in upland Peru and Bolivia, but torrential rains occur along the Pacific coast (Aceituno, 1988; Vuille et al., 1998, 1999; Garreaud & Aceituno, 2001; Vuille & Keimig, 2004).

In contrast, summer rainfall on the southern Altiplano and western Andean flank has a mode of variability that is closely tied to precipitation anomalies and humidity levels over the Chaco region of Argentina (Vuille & Keimig, 2004). Thus, today there are two distinct modes of variability in summer rainfall (Vuille & Keimig, 2004) (Fig. 2). The north-northeast mode is tied to ENSO variability, and the southeast mode is tied to extratropical precipitation anomalies in the lowlands east of the Andes. Unfortunately, a more complete understanding of this southern mode of modern climate in the dry central Andes is hampered by a lack of reliable and continuous precipitation data. Recent advances in the isotope hydroecology of columnar cacti and their spines (English et al., 2007) and tropical dendrochronology (e.g., Evans & Schrag, 2004; Anchukaitis et al., 2008) should produce more detailed records of recent climate throughout South America.

In contrast to the Atlantic moisture falling on the Altiplano and Andes, the Pacific is likely the source of the scant precipitation that falls today in the Atacama. Pacific moisture is effectively excluded from the dry central Andes by the descending limb of the southeast Pacific anticyclone under the influence of the cold Humboldt Current (Vuille, 1999), which has likely been active since the early Tertiary (ca. 65 Ma) (Keller et al., 1997). The steep Coastal Cordillera also limits the inland penetration of Pacific fog to a narrow elevation band (500–1000 m). Although the Coastal Cordillera largely blocks Pacific storms, rare precipitation events may penetrate the Atacama Desert in the austral winter (May through July). These storm fronts typically migrate northward from the westerly precipitation belt that forms the southern boundary of the Atacama at ca. 26°S (Vuille & Ammann, 1997). Today, Pacific sea surface temperature gradients modulate penetration of these Pacific fronts into the Atacama and western Andean flank, and El Niño years are associated with increased precipitation and/or fog intensity in the Atacama (Dillon & Rundel, 1990).

Climate Change in the Dry Central Andes: Mechanisms and Implications

Potential causes of climate change in the dry central Andes include: (1) changes in seasonality, especially local summer insolation (e.g., Baker et al., 2001a, b; Rowe et al., 2002; Fritz et al., 2004); (2)

changes in global temperature (e.g., Blodgett et al., 1997; Garreaud et al., 2003); (3) changes in aridity over the Amazon Basin (e.g., Mourguiart & Ledru, 2003); and (4) changes in sea surface temperature gradients (e.g., Betancourt et al., 2000; Garreaud et al., 2003; Placzek et al., 2006b; Quade et al., 2008). Our lake chronology strongly argues against simple forcing of summer precipitation by summer insolation, and we rule out local January insolation as the primary driver of lake cycles; both deep lakes occur during periods of low to moderate local summer insolation. The Tauca lake cycle reached a maximum between 16.4 and 14.1 ka, ca. 5 ka after the insolation peak at ca. 20 ka (Fig. 3), and the Ouki lake cycle spans the most profound minimum (105–100 ka) in January insolation in the past 200 ka. Similarly, the Ouki lake cycle and the CAPE occur during periods of moderate global temperature, indicating no direct link between precipitation changes and temperature. Past, present, and possibly future climate changes in aridity over the region are, however, likely linked to changes in ENSO variability and moisture level in the eastern lowlands.

CAPE allows examination of the interaction between ENSO and precipitation anomalies over the Gran Chaco lowlands during past wet events over the dry central Andes. Chronology and climate proxy data for CAPE suggest a temporal offset between the Altiplano lake record and the Salar de Punta Negra wetland and rodent midden record. We attribute this to the operation of two separate modes of rainfall over the northern and southern portions of the central Andes during CAPE. The timing of the first phase of CAPE coincides with evidence for intense upwelling (La Niña) in the central Pacific between 18 and 13 ka (Palmer & Pearson, 2003) (Fig. 4). La Niña–like conditions today result in wet years on the Altiplano, and important ancient links may exist between central Andean moisture and Pacific sea surface temperature gradients during the Pleistocene. The modern link between ENSO anomalies and precipitation variability is weaker farther south where CAPE starts ~1000 years later. The second phase of CAPE created the shallower Coipasa lake cycle, but was the more significant precipitation event farther south (Fig. 4). Similarly, modern precipitation anomalies on the western Andean flank to the south are tied more closely to circulation anomalies over the Gran Chaco.

Conclusions

Hyperaridity in the core of the Atacama Desert dominates over a period greater than 10 Ma, in contrast to the western Andean flank and the Altiplano, where evidence from a variety of climate proxies points toward significant changes in paleoprecipitation during the Pleistocene. Over long periods of time (> 10 Ma), the uplift of the Andes and the formation of the Altiplano plateau are critical in the formation of the Andean rain shadow, making the Atacama Desert uniquely long-lived and arid. Conversely, summer insolation over the Altiplano plateau does not appear to drive changes in precipitation over the Altiplano or Amazonia over millennial and glacial-interglacial timescales. Instead, evidence, from both modern climate and paleorecords, increasingly points to ENSO-like variability and extratropical moisture over the Gran Chaco region of Argentina as causal mechanisms for climate variability on the Altiplano and western Andean flank. These two modes of modern central Andean climate variability appear to operate over different geographic regions and at different time intervals. ENSO variability is currently more significant on the Altiplano and, during the earliest phase of CAPE (18.1–14.1 ka), may be linked to intense and prolonged La Niña–like conditions. In contrast, extratropical moisture is today more significant on the western Andean flank and may play a greater role during the latter phase of CAPE (after 13 ka). Thus, modern climate variability and past millennial scale variability appear to be forced by the same mechanisms and suggest that future climate changes in the region will not come as a direct result of temperature shifts, but rather from teleconnections to global circulation patterns such as ENSO.

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